RESEARCH NOTEA comparison between reference transpiration and measurements of evaporation for a riparian grassland site

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RESEARCH NOTE

A comparison between reference transpiration and measurements of evaporation for a riparian grassland site

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Abstract

This paper compares direct measurements of evaporation with the values predicted for reference transpiration. The measurements of actual evaporation were made using an eddy correlation device on a grass field adjacent to the river Thames. Measurements of soil moisture and the driving meteorological variables were also made. The results showed that, during a period with minimal rainfall but no water stress, the cumulative values of reference transpiration compared very well with the cumulative measured evaporation and changes in soil moisture content. However, the values on specific days did not compare well. Following significant rainfall, the measured evaporation increased for a few days, probably due to evaporation of free water from the canopy or soil. Reference transpiration fell consistently below the measured evaporation once the soil moisture deficits exceeded 140 to 150 mm.

Introduction

Most methods of quantifying water resources derive the water balance using relatively simple models of evaporation such as those of Penman (1949) and Monteith, 1965. These models are normally driven with a time series of net radiation, wind speed, air temperature and vapour pressure deficit. In practice, these models are often used to generate estimates of evaporation for a 'reference crop' which is defined as an actively growing crop, fully covering the ground and being well supplied with water (Shuttleworth, 1988). These estimates are then extrapolated to other land cover types by the application of an appropriate crop factor (Doorenbos & Pruitt, 1984). Thus it is critical that the models should estimate accurately the evaporation from the 'reference crop', usually grassland. This paper uses measurements of evaporation, made with the eddy correlation technique to test the model currently recommended by the FAO (Smith, Allen, Monteith, Perrier, Pereira, & Segeren, 1991). The data set includes not only direct measurements of evaporation but also measurements of soil moisture, rainfall and the meteorological driving variables required for the model.

The site

The measurements were made in a meadow near Wallingford (51° 36.1′ N, 1° 6.7′ E.), Fig. 1, on the left bank of the river Thames, at an altitude of about 45 m A.O.D., and which extends across the present flood plain surface and a terrace approximately 1.5 m higher. It has been used as permanent pasture, and occasionally for a hay cut, since shortly after World War II, during which it was cultivated. The vegetation is classified as an *Alopecurus pratensis* variant of the *Lolium perenne-Cynosauriis cristatus* grassland community of the National Vegetation Classification (Rodwell, 1993). A survey carried out at intervals during 1992 established that the leaf area index of the vegetation varied between 0.6 and 3.7. A hay cut was not taken during 1997.

The soils across the meadow drain freely to the water table. They have been mapped as typic-argillic brown earths of the Sutton 2 association (Jarvis, Allen, Fordham, Hazelden, Moffat, & Sturdy, 1984). A layer, about 4.5 m thick, of calcareous gravels and sands with lenses of finer sediment, overlies the Upper Greensand. Over most of the meadow these gravels are covered by 0.7 to 1.5 m of clay loam textured material, but in places remnants of alluvium occur below this deposit. Alongside the river there is a zone about 100 m wide where the soils are developed

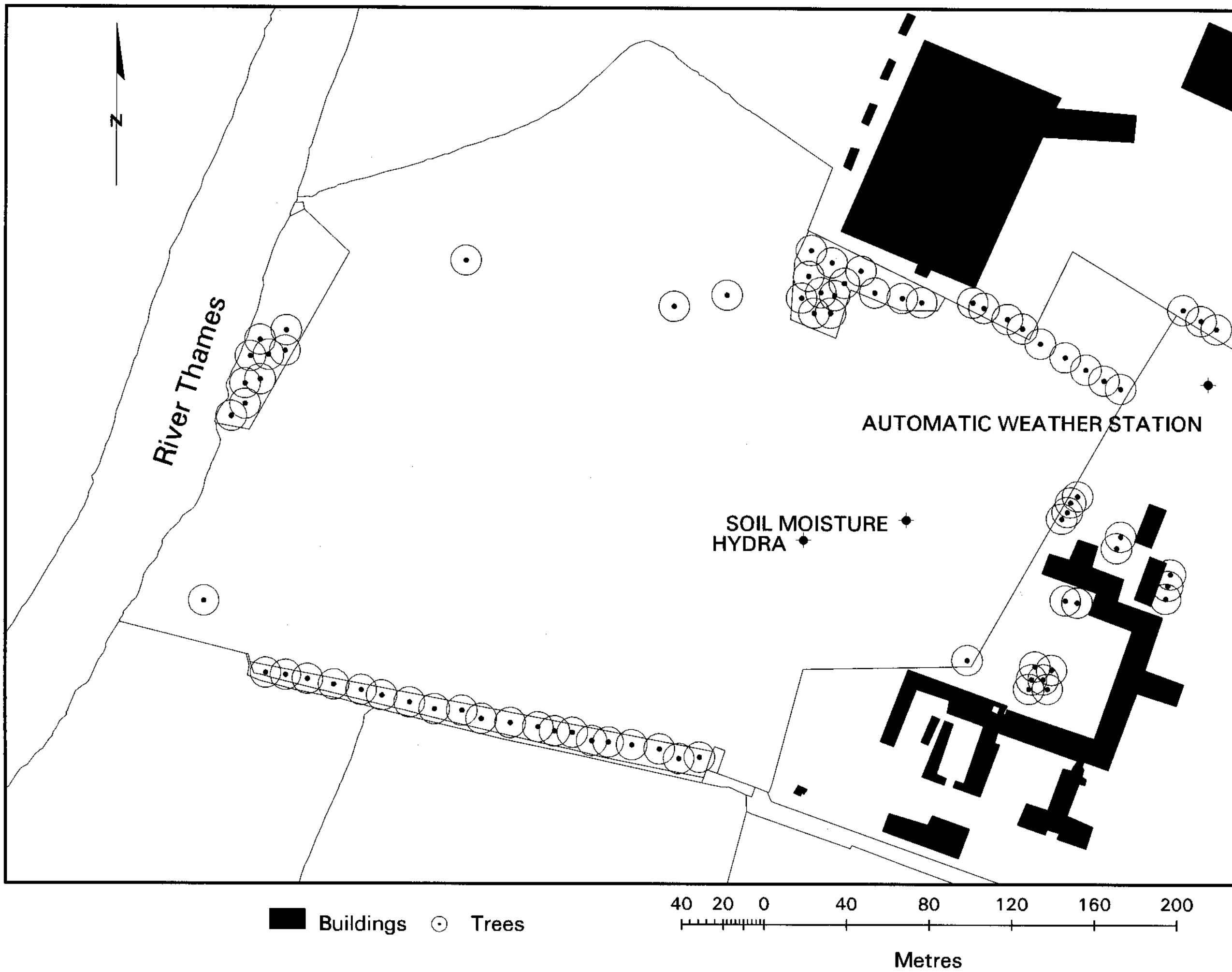

Fig. 1. *Location of the instruments.*

directly in clay loam textured alluvium, 1 to 1.6 m deep. The water table here is influenced directly by the river level which is controlled by a succession of weirs upstream. The water table depth fluctuates about 1.4 m below ground level, rising in winter so that occasionally this part of the meadow floods. The water table is deeper below the terrace surface, fluctuating between 1.5 and 2.5 m below ground level. Throughout 1997, no flooding occurred and the water levels remained below 2.5 m at the terrace site.

The total rainfall at Wallingford for 1997 was 518.8 mm which is below the mean annual rainfall of 583 (±83) mm for the period 1962 to 1997. The rainfall was highly variable through the year. A dry January was followed by a very wet February with rain falling on almost every day (Fig. 2). Very little rain fell during March and the first half of April but was followed by heavy rainfall through late April and the first half of May. Through the following summer there was a tendency for wet periods of 7 to 10 days to be followed by relatively dry periods of 10 to 20 days. Heavy rainfall characterized the last two months of the year. The extended dry period in March and April and the sequence of wet and dry periods provide a contrasting set of conditions with which to study the factors affecting evaporation.

Measurements

Soil water content was measured by a neutron probe (Bell, 1987) in an access tube at intervals of 0.1 m down to a depth of 1 m., and then at intervals of 0.2 m. to the bottom of the tube at a depth of 2.9 m. An array of mercury manometer tensiometers was also installed, at the same location, at intervals of 0.1 m down to a depth of 0.8 m. and then at intervals of 0.2 m. down a depth of 2.6 m. Readings were taken at intervals of about 7 days.

Direct measurements of evaporation were made using a Mk. 2 Hydra eddy correlation system (Shuttleworth *et al.*, 1988). This system measures the evaporation and sensible heat and momentum fluxes which are stored as mean val-

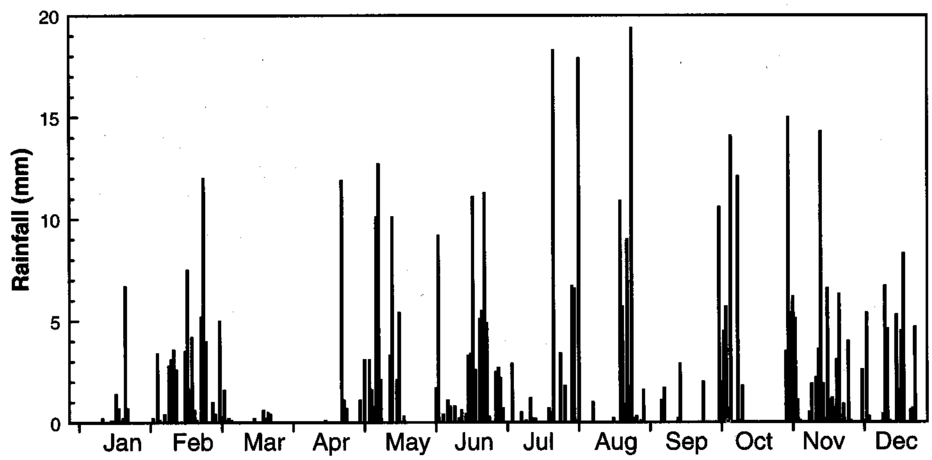

Fig. 2. *1997 daily rainfall at the site.*

ues. There are inevitably gaps in the data due to rain on the sensors, rapid changes of temperature and humidity, and sensor and logger failures. The Hydra logger software automatically detects when the data are unreliable and only data which were coded without fault, about 70% of the total, have been included in the analysis below. While the uninterrupted fetch over the meadow is not ideal for micrometeorological measurements (80 m to 350 m depending on the wind direction), the successful energy closure, Fig. 3, suggests that the measurements are well founded.

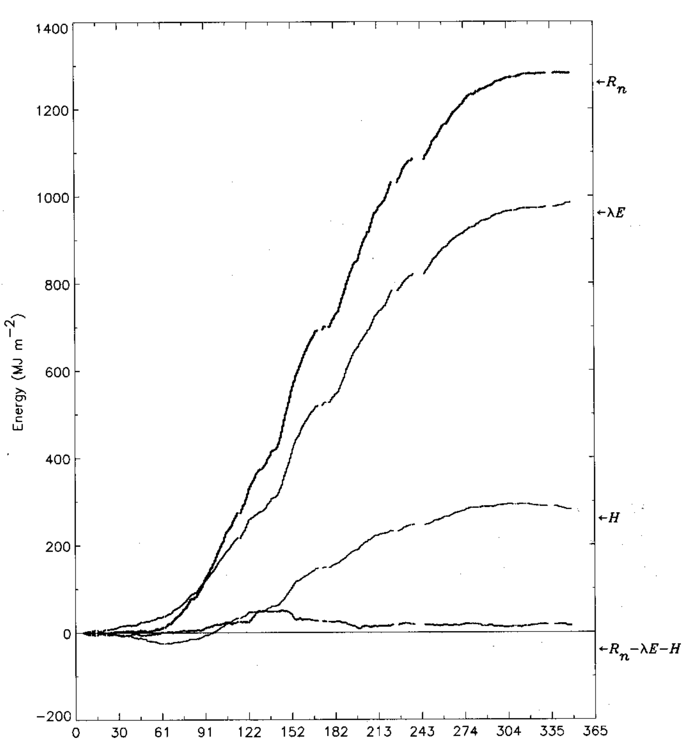

Fig. 3. *Cumulative fluxes and energy closure.*

The cumulative measured energy balance components data for 1997 are shown in Fig. 3. It should be noted that these are not totals for the entire year but are the hours when all three fluxes (latent heat flux, λE, sensible heat flux, H and net radiation, R_n) are available. The figure shows the partition of energy and the energy closure. Over the year, between 80 and 90% of the radiation goes into latent heat flux; this proportion is slightly less during the summer whilst in the winter the evaporation can exceed the net radiation. The energy closure is excellent and the net radiation is within 5% of the sum of the turbulent fluxes. There is, as expected, some variation through the year due to changes in soil heat storage.

The Hydra also records humidity, air temperature, windspeed and net radiation. Humidity is measured with a RH sensor (Vaisala, Finland), the air temperature with a fine thermocouple sensor used for the measurement of temperature fluctuations and the net radiation with a radiometer (REBS Q*6, Seattle, Wa, USA). To provide the continuous record of driving data required to calculate reference transpiration, these data were supplemented by data from the automatic weather station (AWS) operated at a site 200 m NE of the Hydra. Comparisons showed that the agreement between the data taken by the Hydra and the AWS is excellent.

Deriving daily totals from the hourly measurements of fluxes

The hourly total flux values measured by the Hydra were aggregated to give daily total values. These values were compared with estimates of reference transpiration calculated using daily values of the driving variables. The mean daily fluxes were calculated using the following strategy:

* all hourly values not coded as error free were rejected;
* when the latent heat flux measurement was missing it was set to zero when the net radiation was less than 10 $W\ m^{-2}$;

* any days where all the flux values for more than half the daylight hours were missing were rejected;
* for the remaining days, where an hourly value of the latent heat flux was missing it was estimated from the hourly net radiation assuming the same Bowen ratio ($H/\lambda E$) as that of the next hour with measured fluxes;

Where a daily value of the latent heat flux was missing it was estimated from the daily net radiation by assuming the same Bowen ratio as the next day with a measured value so long as they were within the following few days.

This resulted in 263 days for which no hourly values had to be estimated, 46 days when one or more hourly value had to be estimated, 6 full days which were interpolated and 50 days for which there were no data.

Calculation of reference transpiration

Cain *et al.* (1998) recommend the use of the term reference transpiration to indicate the evaporation calculated using the Penman-Monteith model according to Allen *et al.* (1994), with a constant surface resistance of 70 sm^{-1}, a constant vegetation height of 0.12 m, and daily meteorological data viz:

$$\lambda E = \frac{\Delta(R_n - G) + 86.4\rho c_p \frac{(e_a - e_d)}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \quad (1)$$

where:

E reference transpiration ($kg\ m^{-2}\ d^{-1}$)
G soil heat flux ($MJ\ m^{-2}\ d^{-1}$)
R_n net radiation flux density at the surface ($MJ\ m^{-2}\ d^{-1}$)
c_p specific heat of moist air ($kJ\ kg^{-1}\ °C^{-1}$)
e_a saturation vapour pressure (kPa)
e_d saturation vapour pressure computed at dew point (kPa)
r_a aerodynamic resistance ($s\ m^{-1}$)
r_s bulk surface resistance of the vegetation canopy ($s\ m^{-1}$)
Δ saturation vapour pressure curve slope ($kPa\ °C^{-1}$)
λ latent heat of vaporisation ($MJ\ kg^{-1}$)
γ psychrometric constant ($kPa\ °C^{-1}$)
ρ density of air ($kg\ m^{-3}$)

The aerodynamic resistance is given by:

$$r_a = \frac{\ln\left(\frac{z_m - d}{z_{om}}\right)\ln\left(\frac{z_h - d}{z_{oh}}\right)}{k^2 U_z} \quad (2)$$

where:

U_z mean wind speed at height z ($m\ s^{-1}$)
d zero plane displacement of wind profile (m)
k von Karman constant
z_h height of air temperature and humidity measurements (m)
z_m height of wind speed measurement (m)
z_{oh} roughness parameter for heat and water vapour (m)
z_{om} roughness parameter for momentum (m)

d is taken to be $2/3h_c$, where h_c is the mean vegetation height which is taken as a constant value of 0.12 m. The roughness lengths are given by:

$$z_{om} = 0.123h_c \quad (3)$$

$$z_{oh} = 0.1z_{om} \quad (4)$$

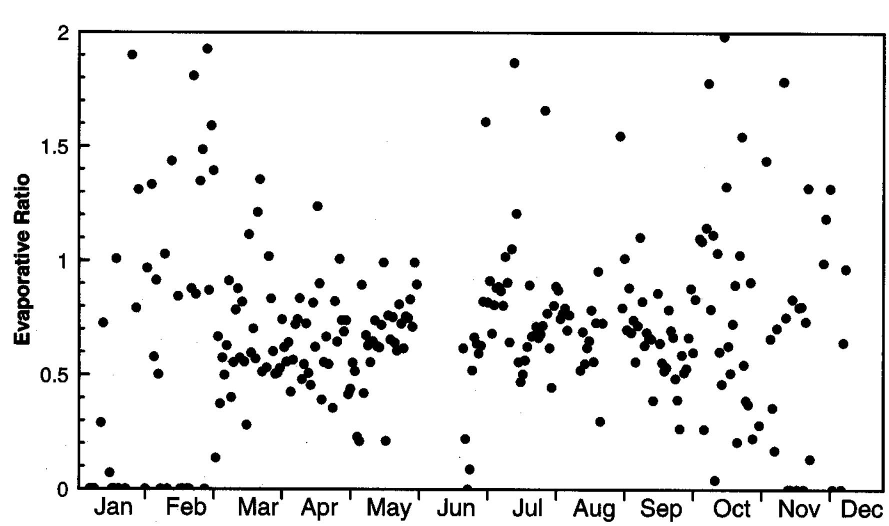

Fig. 4. *Daily evaporative ratio, $\lambda E/Rn$.*

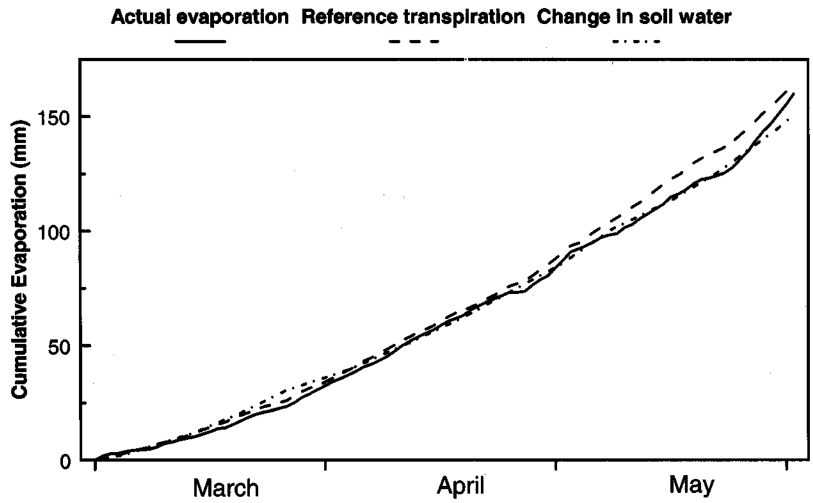

Fig. 5. *Comparison of cumulative actual evaporation, reference transpiration and change in soil water during a period when evaporation was not limited by soil water availability.*

Results

The variation in the evaporative ratio, $\lambda E/R_n$, is shown in Fig. 4. Both at the beginning and the end of the year, the evaporative ratio consistently exceeds unity implying large scale advection of energy with the evaporation being driven by the wind speed and humidity. During the summer months, the evaporative ratio drops to a value of around 0.7.

Figure 5 shows the cumulative measured evaporation and reference transpiration for March, April and May as evaporation becomes increasingly a function of net radiation but before the soil moisture deficit has reached a point which limits transpiration. During this period, reference transpiration agrees well with the measured evaporation.

Figure 5 also shows the cumulative change in the soil water content derived from the neutron probe measurements. The soil water content was taken as the total water content from the surface to a depth of 1.6 m; data from the tensiometers showed that this was the maximum depth reached by the zero flux plane and hence defined the zone affected by evaporative losses (Wellings, 1984). A sequence of the daily change in the soil water content was generated by linearly interpolating the water content between dates when readings were taken and then subtracting the rainfall. The result shows excellent agreement with the evaporation measured using the eddy correlation technique and reference transpiration; this confirms that the measured values of daily evaporation are accurate and that the reference transpiration simulates these well.

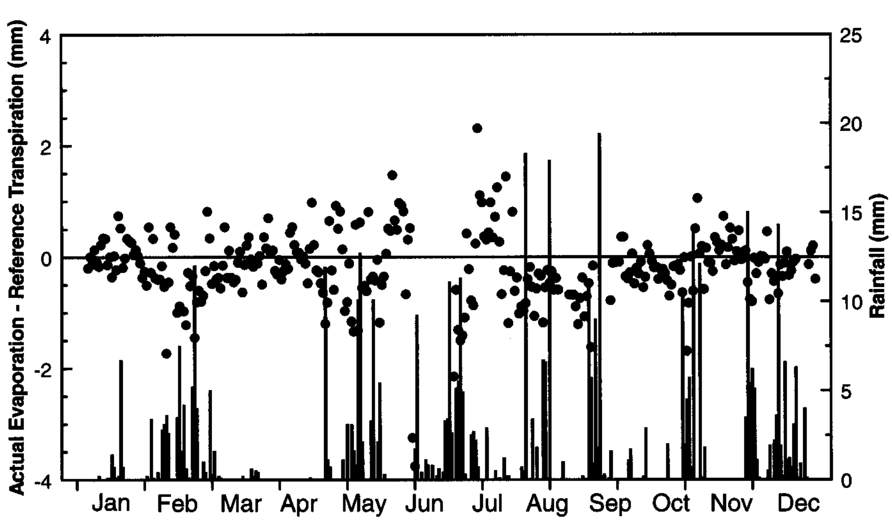

Fig. 6. *Daily differences between actual evaporation and reference transpiration.*

Examination of the daily values of measured evaporation and reference transpiration shows that, although the values aggregated over periods of many days agree well, agreement between values on specific days is poor. Figure 6 shows the time sequence of the difference between measured evaporation and reference transpiration. The values covering March, April and May cluster around the zero line but there is a large scatter about this line; the average difference is 0.05 with a standard deviation is 0.56 $mm\ d^{-1}$.

There is a tendency for the measured evaporation to exceed reference transpiration for several days following periods of prolonged rainfall, for example in mid–May and late June, and in a prolonged period through much of October. This may be a result of evaporation from the free water in the soil, the canopy and the accumulation of dead foliage beneath the canopy or a combination of these.

There is a short period in late May–early June when reference transpiration consistently exceeds the measured evaporation; this suggests that transpiration was limited by the low soil water content. Following heavy rainfall in mid-June, evaporation increased but a second period of evaporation below the reference value began in the first week of July and was sustained until the beginning of October. Heavy rainfall, together with decreasing evaporative demand as a result of decreasing net radiation, then restored the soil water content to a level that could support the full capacity for transpiration by the vegetation. This interpretation is confirmed by comparing the soil moisture deficits predicted by a simple water balance model, calculated using the reference transpiration, with the observed soil moisture deficits (Fig. 7).

The simple water balance model used was:

$$\theta_i = \theta_{i-1} + P_i - E_i \qquad (6)$$

where

E_i reference evaporation on day i (mm)
P_i rainfall on day i (mm)
θ_i soil water content on day i (mm)
θ_{i-1} soil water content on day i–1 (mm)

The soil moisture deficits were then calculated as the difference between the soil water content and the soil water content at field capacity, 453 mm, which was determined from measurements of soil water content and potential as the soil water content at a potential of –10 kPa (Marshall, Holmes and Rose, 1996).

The modelled and observed soil moisture deficits show very good agreement from the beginning of the year until mid-May when the model begins to over-estimate the deficits. This corresponds to the beginning of the first period when the measured evaporation consistently fell below the reference transpiration. The difference between the modelled and the observed values increases again in early July, a trend which continues until the beginning of October, reaching a maximum of 73 mm. This again corresponds to the period when the reference transpiration exceeded the measured evaporation. The soil moisture deficits which occur when the measured evaporation began to fall below the reference transpiration are between 140 and 150 mm.

Prior to the adoption of the Penman-Monteith model of evaporation, the model of Penman (1949) was used to calculate potential evaporation and it is interesting to make a comparison of the cumulative evaporation predicted by the two models (Fig. 8). Although the annual totals predicted by the two models are almost identical, the Penman model predicts higher daily values during the winter but lower during the summer—in agreement with the theoretical analysis of Thom and Oliver (1987). The values from the Penman-Monteith model follow the change in soil moisture, prior to the onset of stress, more closely than those

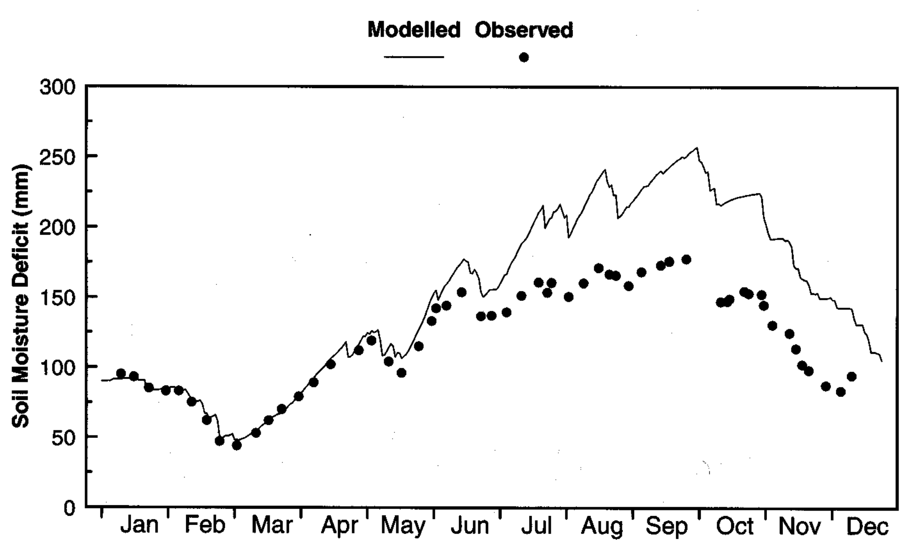

Fig. 7. *Comparison between observed soil moisture deficits and the values predicted using a simple model.*

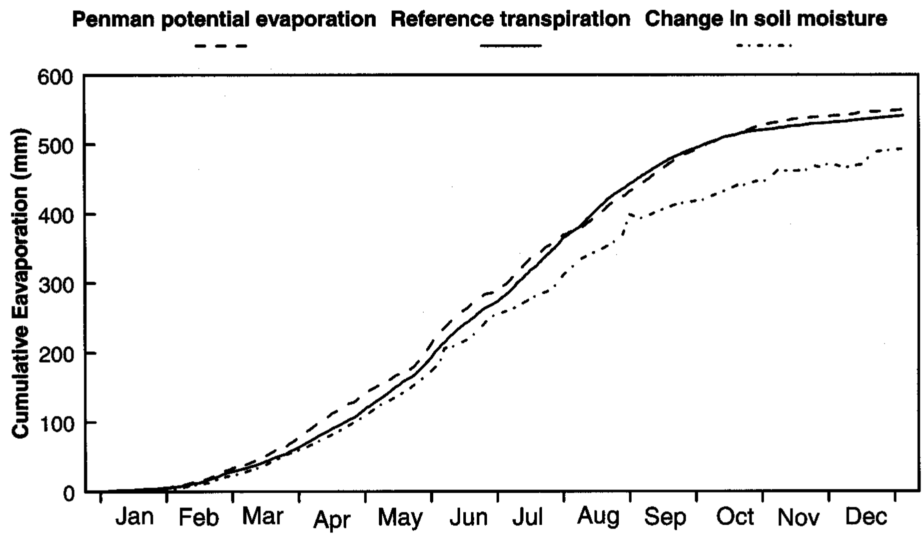

Fig. 8. *Comparison of cumulative Penman potential evaporation, reference transpiration and change in soil moisture.*

from the Penman model. This confirms that the Penman-Monteith model will result in more accurate estimates of potential evaporation.

Conclusions

The direct measurements by eddy correlation show a large proportion of the incoming energy (about 80%) was used in evaporation through the entire year. During the summer months this proportion dropped to below 70%. However, during the winter months when net radiation was low, wind speed, air temperature and vapour pressure deficit often became the dominant meteorological driving variables and the evaporation frequently exceeded the radiation input.

The simple, daily Penman-Monteith model, with a constant bulk surface resistance of 70 $s\ m^{-1}$, simulated the cumulative evaporation correctly, in the absence of stress due to soil moisture deficits. However, its ability to simulate specific daily values is poor, most likely because a single value of surface resistance takes no account of its variation due to factors, such as temperature, vapour pressure deficit and incoming radiation (Stewart & Verma, 1992). Under these conditions, the simple model proposed by Allen *et al.* (1994) will be more useful to hydrologists, who are mostly interested in seasonal patterns and totals of water use, than to meteorologists who require a description of the changes in the fluxes on an hourly and daily time scale.

Rainfall at this site during 1997 was well distributed and there was no 'summer drought' situation, nevertheless, significant soil moisture deficits developed and the effect of stress due to limited soil water content was observed from mid-May onwards. Evaporation was reduced below the reference transpiration rate once a soil moisture deficit of between 140 and 150 mm was exceeded. After rainfall, evaporation rose above the reference transpiration for at least 10 days before returning to a stressed value. After substantial rainfall at the end of September, the evaporation reverted to the reference transpiration value, apparently not affected by water stress although a substantial deficit existed through the entire soil column. It is evident that at this time water extraction from the roots can occur from the wetted surface layers. Such patterns cannot be simulated by simple soil water models and require a multilayer soil model in which the stress function can respond quickly to the presence of water in the upper soil layers.

Acknowledgements

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